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Performance of an Iterative Analysis-Initialization Scheme at Low Latitudes in NOGAPS

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In common with many atmospheric models, the Navy Operational Global Atmospheric Prediction System (NOGAPS) tends to reject low-latitude height and divergent-wind data, thus the information added to the analysis by this data is largely removed during initialization. This study addresses this problem by repeatedly iterating the multivariate optimum interpolation analysis and nonlinear normal mode initialization of NOGAPS, with the goal of obtaining a balanced initial state in reasonable agreement with the original analysis. The impact of various optimum interpolation parameters on the results is examined; these parameters determine such analysis properties as mass-wind balance, horizontal divergence, and the length scale over which observations influence grid points. Results show minimal impact of the iterative scheme on data rejection, for all selected values of analysis parameters. There is little improvement in data retention when the background error variance is decreased, although the individual analysis- and initialization-produced changes become smaller. Diabatic normal mode initialization produces only marginal positive impact on data acceptance, because the initialized vertical modes are nearly barotropic in the troposphere, thus only weakly influenced by diabatic forcing.

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1. Introduction

Analysis and initialization have traditionally been regarded as two separate components of an atmospheric prediction system. While this is gradually changing with the advent of more unified approaches (e.g. Parrish and Derber, 1992), many operational systems still employ a two-step technique consisting of optimum interpolation (Gandin, 1963) and nonlinear normal mode initialization (Machenauer, 1977). This method has achieved considerable success in practice, but suffers nevertheless from a fundamental incompatibility between its two components. In particular, optimum interpolation is designed to produce an analysis fitting the data as closely as possible, taking into account the presumed error characteristics of both the observations and a first guess. Nonlinear normal mode initialization, however (and for the most part initialization in general), is typically applied after the completion of the analysis with the sole aim of eliminating high-frequency oscillations from the ensuing forecast. It is (usually) not constrained by data quality but rather by the rotational mode component of the analysis. As a result, large changes may be made to the analysis even in regions where the data is highly accurate, if that data projects most strongly onto gravitational modes (Daley, 1980). This is an undesirable instance of the phenomenon of data rejection, in which information added during the analysis phase is removed by the initialization or forecast model.

To minimize the rejection of accurate data while still producing a dynamically balanced state, some investigators have adopted a variational approach. For example, Daley (1978) and Tribbia (1982) employed a technique which sought to minimize changes to an analysis under the constraint of nonlinear normal mode balance. This method, however, proved computationally intractable for realistically truncated models. Fillion (1991) and Fillion and Roch (1992) developed a scheme known as variational implicit normal mode initialization, in which the balance condition was applied in physical, rather than normal mode, space. This substantially reduced the computational requirements and allowed implementation at much higher model resolutions. Results with the method have generally appeared promising, although it should be noted that the imposed balance is not identical to that of the standard

nonlinear normal mode technique; see Fillion (1991) and Temperton (1989) for further elaboration. The common feature of all such variational schemes is that they minimize, subject to some dynamical constraint, the difference between an initialized state and the original analysis.

An alternate method of combining analysis and initialization is to attempt to minimize the difference between an initialized field and the *observations* (rather than the analysis), as in Williamson and Daley (1983). Their technique was to first perform an analysis and initialization in the usual manner, then use the initialized analysis as the background field for a reanalysis employing the same observations. (The background state for optimum interpolation normally consists of a short-range model forecast.) The reanalysis was initialized and the process repeated iteratively until the changes produced by both analysis and initialization decreased to less than some desired value. Williamson and Daley (1983) successfully used this scheme to mitigate data rejection in strongly curved midlatitude flow. This rejection stemmed from the geostrophic coupling assumed by optimum interpolation, which leads to inappropriate balance in regions where the ageostrophic component of the gradient wind is significant (see Williamson et al., 1981). Application of the iterative method yielded a balanced state which differed everywhere from the original analysis by an amount compatible with expected analysis error.

Although Williamson and Daley (1983) (and many other studies) considered data rejection only in midlatitudes, this phenomenon is perhaps an even greater problem in the tropics. There, the smallness of the Coriolis parameter results in an effective decoupling of the mass and wind fields, at least from the standpoint of optimum interpolation which allows only a geostrophic-type balance. In addition, conventional observations at low latitudes are extremely sparse, and thus their impact on the analysis tends to be limited to isolated regions which are separated by data-void areas. This negatively affects the dynamical balance of the analysis as discussed in Williamson et al. (1981). Certain types of data are most likely to be rejected at low latitudes; in particular, the generally large Rossby deformation radius (Rossby, 1938) means that there is a strong tendency for the unbalanced height field to adjust to the wind. Any mass-wind imbalance produced by the analysis will therefore be removed through rejection of the analyzed height, provided the vertical scale of

the imbalance is sufficiently large (the Rossby deformation radius is proportional to internal gravity wave speed). Divergent wind data also tends to be rejected, as according to the geostrophic adjustment theory of Rossby (1938), the initial divergent wind projects only onto gravity-wave modes. This occurs regardless of the value of the Rossby deformation radius and so the rejection of divergence is typically found at all latitudes. (An exception is for divergence forced predominantly by latent heating or other diabatic processes, as long as these processes are included in the initialization; we will return to this point in Section 3). A discussion of the retention of different types of data in the context of normal mode initialization appears in Daley (1980).

Errico et al. (1993) provide a demonstration of low-latitude data rejection in an operational forecast model. The model employed was the Navy Operational Global Atmospheric Prediction System (NOGAPS), a global spectral model which uses optimum interpolation and nonlinear normal mode initialization to define the initial state. Results showed a strong tendency for the rejection of tropical height; the rejection appeared most prominently in isolated centers around low-latitude radiosonde observations, where the data had produced the greatest impact on the analysis. The divergence was largely rejected at all latitudes, consistent with previously discussed theoretical arguments. (Interestingly, the rejection of the divergent wind component was much more apparent in velocity potential than in the divergence itself.) Similar results have been observed in other forecast models; for example, the National Meteorological Center's operational global model (prior to implementation of spectral statistical interpolation) exhibited a pattern of tropical height rejection comparable to that of NOGAPS (see Figs. 4 and 9 in Parrish and Derber, 1992.)

To investigate whether the iterative technique of Williamson and Daley (1983) was a viable method of overcoming low-latitude data rejection, Van Tuyl (1995) applied their scheme to a global, spectral shallow-water model. An analysis from NOGAPS was projected onto the NOGAPS external vertical mode; it served as the "true" atmospheric state for a series of experiments. Using "data" interpolated from this analysis and an initial background field (for optimum interpolation) obtained from a shallow-water model forecast, Van Tuyl (1995) iterated Williamson and Daley's scheme for a variety of analysis parameter values. (These parameters control such aspects of optimum interpolation as the degree of

mass-wind coupling, the amount of divergence, and the background error correlation length scale.) For every case examined, the iterative method was able to achieve only very modest improvement in tropical data retention; this result was insensitive to the values of any of the analysis parameters. It had originally been hypothesized that the rejection was due to an inappropriate tropical value of one or more parameters, especially the parameter coupling the analyzed mass and wind. This was discovered not to be true, however, at least for the particular model and analysis technique employed.

The results of Van Tuyl (1995) were obtained using a barotropic model with no diabatic forcing. To determine the applicability of these results to a more realistic model, the present study examines the performance of the Williamson and Daley (1983) scheme in NOGAPS, the same model employed by Errico et al. (1993). As in Van Tuyl (1995), particular attention is given to the sensitivity (or insensitivity) of the method to analysis parameter values in the tropics. Although NOGAPS includes an extensive variety of diabatic processes, its initialization is adiabatic and so differences with the shallow-water model results will primarily reflect the presence of vertical structure in NOGAPS. (The use of actual data, rather than synthetic observations as in Van Tuyl, 1995, may also have an impact.) An option for diabatic normal mode initialization of NOGAPS does exist, however, and in section 3 we briefly examine its influence on data rejection.

A short description of NOGAPS, concentrating on its analysis and initialization components, appears in Section 2. Section 3 presents the results of experiments with the iterative analysis-initialization scheme. The influence of selected optimum interpolation parameters on these results is examined. The summary and conclusions are given in Section 4.

2. Model Description and Analysis Method

a. NOGAPS

NOGAPS 3.3, the version of NOGAPS employed in this study, is a global, spectral model with 18 sigma-coordinate levels. The horizontal truncation for our experiments is T79, corresponding to a 1.5-degree resolution in latitude and longitude. (The operational version of NOGAPS currently has truncation T159.) A number of physical processes are included in the model, some of the more significant being long- and short-wave radiation,

stable precipitation, subgrid-scale vertical fluxes, shallow cumulus mixing, gravity-wave drag, and parameterized convection. Optimum interpolation and nonlinear normal mode initialization are employed to obtain the initial state; these techniques are discussed in greater detail in Sections 2b and 2c, respectively. A thorough description of NOGAPS 3.2, which is generally quite similar to NOGAPS 3.3, appears in Hogan and Rosmond (1991).

b. Optimum Interpolation

The optimum interpolation scheme used for NOGAPS is discussed in detail by Goerss and Phoebus (1992). It is a tropospheric analysis performed on 16 pressure surfaces, ranging from 1000 to 10 mb. All except the 925 mb surface are located at standard pressure levels. Data sources include radiosondes, pilot balloons (pibals), surface observations (including buoys and ships), aircraft, and satellites. The satellite data comprises cloud-tracked winds, temperature soundings, and SSM/I wind speeds. Synthetic observations are also created for certain applications, such as bogusing of tropical cyclones; refer to Goerss and Phoebus (1992) for further discussion. The first guess (background) field for the analysis consists of a 6-hour NOGAPS forecast run from the previous initialized analysis; this process is repeated as an intermittent data-assimilation cycle. Quality control of observations is another important feature of the NOGAPS analysis and is described by Baker (1992).

The analysis is multivariate at all latitudes, meaning that each observed variable (u, v, and height) is allowed to influence all analyzed variables. In particular, the analyzed mass and wind are coupled, and the observed height influences the analyzed wind (and the observed wind the analyzed height). However, the degree of this coupling is a function of latitude, with the mass-wind constraint relaxed equatorward of approximately 20 degrees. The result is an analysis in which the *increments* (defined as analysis minus background field) are nearly geostrophically-balanced in midlatitudes, but only weakly balanced in the tropics (that is, the mass and wind are effectively decoupled). The amount of geostrophy in the analysis increment is controlled by the parameter μ; its value in NOGAPS is 0.9 in midlatitudes and 0.5 at low latitudes (a value of 1 corresponds to exact geostrophic balance). A more detailed discussion of this parameter may be found in Lorenc (1981) and Daley (1985). The analyzed u and v are similarly coupled through a divergence parameter,

ν (Daley, 1985), which is 0 (1) for nondivergent (purely divergent) flow. The value of ν in NOGAPS is 0 in midlatitudes and 0.1 in the tropics; thus, the midlatitude analysis increments contain no horizontal divergence, whereas the divergence at low latitudes may be significant (the response of divergence to ν is highly nonlinear; see Daley, 1985, and Unden, 1989). The background error correlation is approximated by a modified second-order autoregressive (SOAR) function (Franke et al., 1988), with length scale 385 km and nondimensional asymptote 0.1. Further details on the NOGAPS analysis parameters and their theoretical significance are given by Barker (1992).

Due to computational restraints, it is not feasible to analyze all of the available data simultaneously; therefore, the volume method suggested by Lorenc (1981) is employed. An optimum interpolation is performed separately for each of 838 volumes spanning the depth of the troposphere. These volumes vary in horizontal extent such that each contains a comparable number of observations; the number of analysis points per volume is additionally influenced by the convergence of the meridians. (For example, the polar cap volumes contain an especially large number of grid points.) Within an analysis volume, all observations are allowed to influence the analysis at each grid point. To minimize discontinuities at the boundaries of volumes, the volumes overlap such that most grid points (specifically, all except those in the polar caps) are contained in four analyses. The overlapping analyses are merged using spatially-dependent weighting functions as described in Goerss and Phoebus (1992).

c. Nonlinear Normal Mode Initialization

Two iterations of the nonlinear normal mode technique of Machenauer (1977) are applied to the completed analysis. The version used by NOGAPS is adiabatic and includes only the three deepest vertical modes; this helps to ensure rapid convergence of the scheme (Ballish, 1981; Errico, 1983). Because the initialized modes are not those typically influenced by diabatic forcing (Errico et al. 1993), one would expect the use of diabatic normal mode initialization to have little additional impact. (This hypothesis is tested in section 3.) The rotational mode coefficients are held fixed at their original values (i.e., the values obtained from the analysis) during initialization, which is the constraint typically employed

in the Machenauer scheme. The combined analysis-initialization technique, therefore, will be successful only to the extent that the slow-mode component of consecutive analyses is appreciably modified. This, in turn, requires that the rejected data initially possess *some* projection onto rotational modes, otherwise that data will have no impact no matter how many iterations are performed.

3. Results

a. Control Experiment

The control experiment consists of six iterations of the combined analysis-initialization scheme beginning from the operational analysis valid 0000 UTC 17 June 1995. The analysis parameters have values nearly identical (in some cases exactly identical) to those of the control experiment in Van Tuyl (1995). Specifically, the geostrophic and divergent coupling parameters μ and ν are respectively 0.9 and 0 in midlatitudes, and 0.5 and 0.1 in the tropics, for both studies. The background error correlation length scale is 385 km in the present experiment versus 417 km in Van Tuyl (1995). The nondimensional asymptote of the correlation (SOAR) function in NOGAPS is 0.1; dimensionally, the error covariance at large distances approaches 10 percent of the covariance at zero distance (i.e., the variance). The corresponding value in Van Tuyl (1995) is identically zero, meaning there is effectively no error covariance at large distance. All analysis parameters, in particular those defining the background error field, are held fixed throughout the iterative scheme.

Results were examined for the 850, 500, and 200 mb levels. (The NOGAPS model employs a sigma, rather than pressure, vertical coordinate, so vertical interpolation using cubic splines is performed as necessary.) In this study we concentrate mainly on the results at 500 mb, since this level is, to a first approximation, most representative of the troposphere as a whole. Results for other levels are generally similar, but one feature of the NOGAPS analysis and initialization schemes regarding vertical dependence should be noted. The analysis produces changes to the background field (analysis increments) which may have any vertical structure, depending on the nature of the data and the background state employed. The NOGAPS initialization, however, includes only the three deepest vertical modes, so changes made to the analysis by this scheme (the initialization increments) will be nearly

constant with height in the troposphere (Van Tuyl, 1995). In particular, the data rejection must be constant with height and proportional to the projection of the analysis increments onto the initialized modes. If an analysis increment is concentrated near one vertical level, the rejection will be "spread" vertically to other levels where no change was produced by the analysis; thus, assessing data rejection by comparing analysis and initialization increments at a given level may be misleading. Also, since the initialized state in our scheme serves as the background for the next analysis, such initialization-produced changes may result in large analysis increments appearing at levels where they were originally negligible.

Figures 1a,b respectively show the first analysis and initialization increments for the 500 mb height field. Note that the initialization increment is substantially larger than the analysis increment for the feature northeast of New Guinea; this reflects the approximately barotropic nature of the initialization increments as discussed in the preceding paragraph. The analysis increment for this feature has minimum amplitude at 500 mb (relative to the other two levels examined), whereas the initialization increment has at all levels a magnitude comparable to that of the vertically averaged analysis increment. Other low-latitude features, in particular those over Brazil and in the central Pacific near 10S,140W, have analysis and initialization increments of comparable magnitude at 500 mb. There is a strong tendency for cancellation between virtually all the tropical height analysis and initialization increments, consistent with the results of Van Tuyl (1995) and Errico et al (1993).

Figures 1c,d present the analysis and initialization increments, again at the first iteration, for 500 mb velocity potential. (The streamfunction field, for reasons discussed in Van Tuyl, 1995, is not shown.) Cancellation between the two increment fields is apparent at all latitudes, being especially well-defined over South America; this global cancellation is also consistent with the findings of Van Tuyl (1995) and Errico et al (1993). As for the height field, the analysis and initialization increments at a single level do not always exhibit a clear cancellation in response to data rejection.

Six iterations of the combined scheme were performed following Van Tuyl (1995). The increments at the conclusion of this procedure are shown in Fig. 2. The 500 mb height analysis increment (Fig. 2a) has significantly decreased since iteration 1 in midlatitudes,

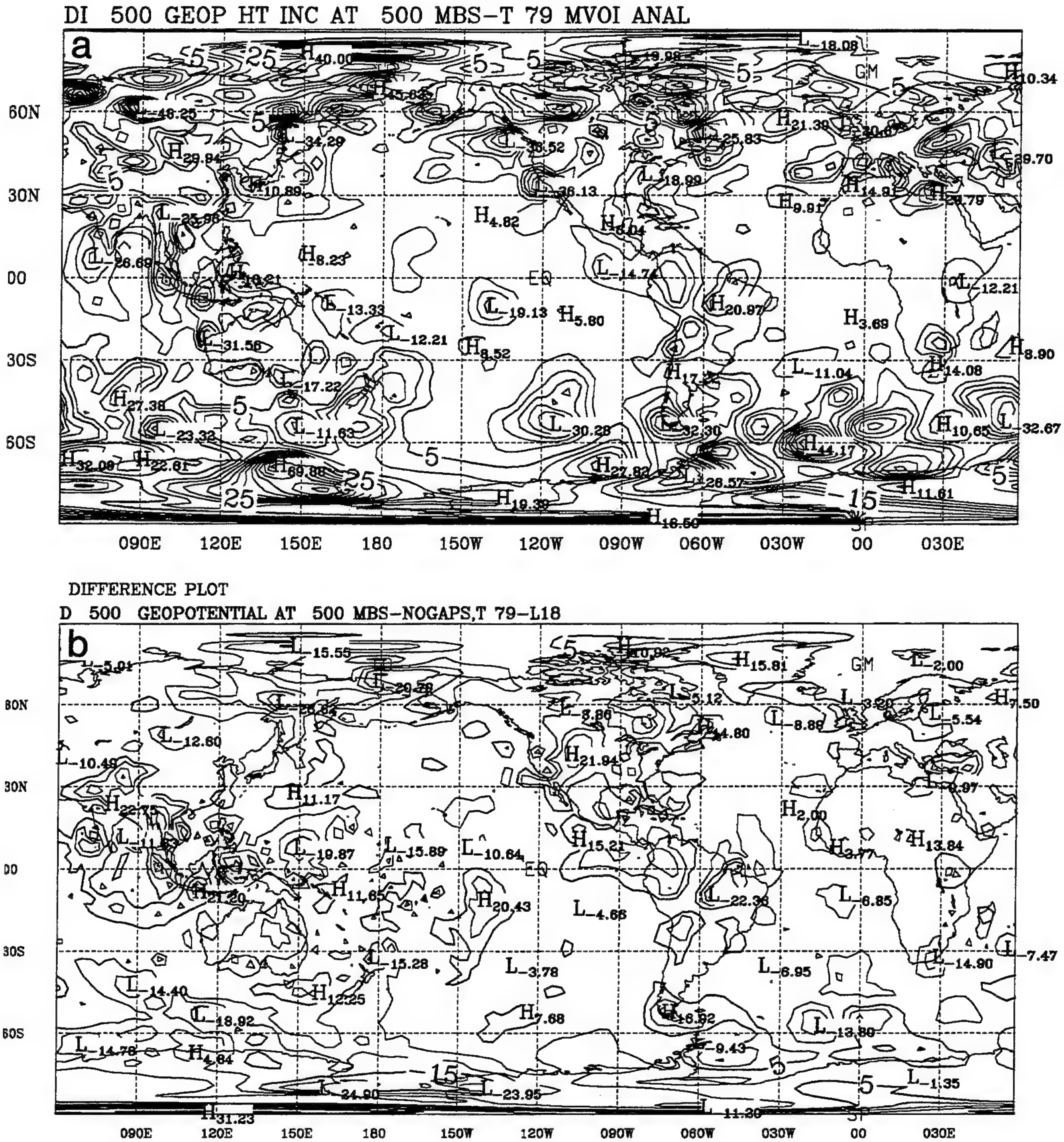

Figure 1. The 500-mb (a) height analysis increment, (b) height initialization increment, (c) velocity potential analysis increment, (d) velocity potential initialization increment at iteration 1 of analysis-initialization scheme for 0000 UTC 17 June 1995 (control experiment). Contour intervals are (a) and (b) 5 m, (c) and (d) 2×10^5 m^2 s^{-1}. The zero contour is omitted.

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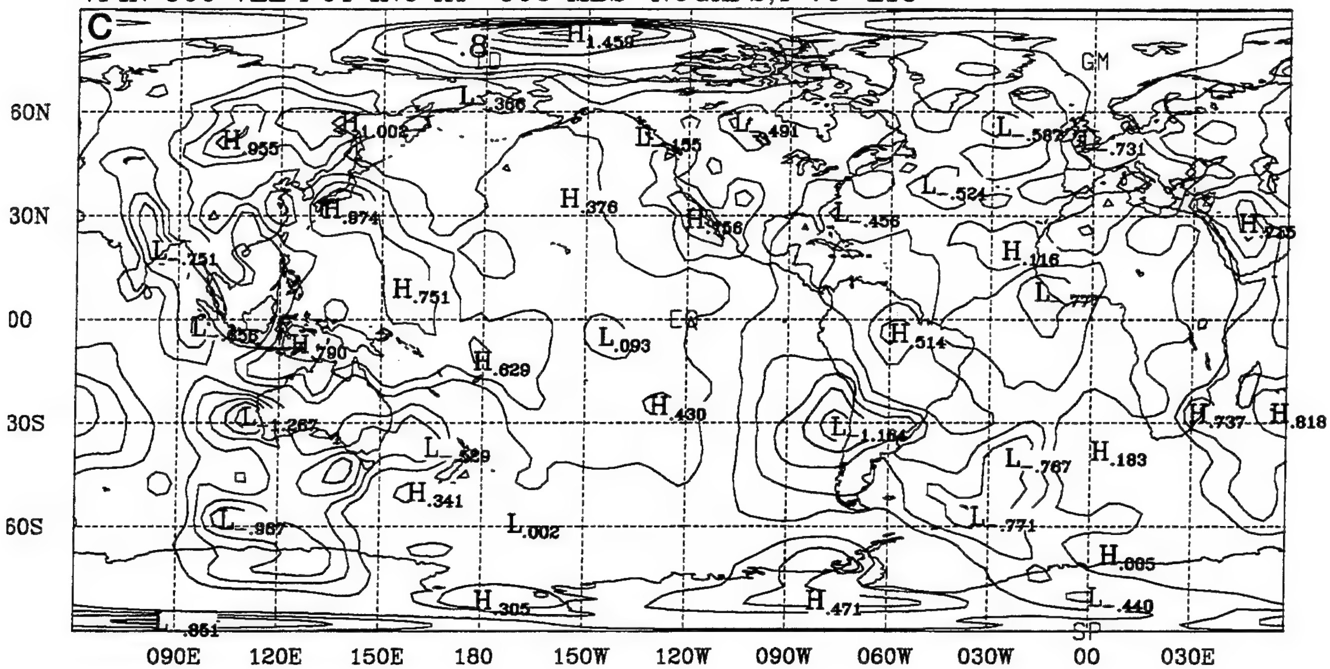

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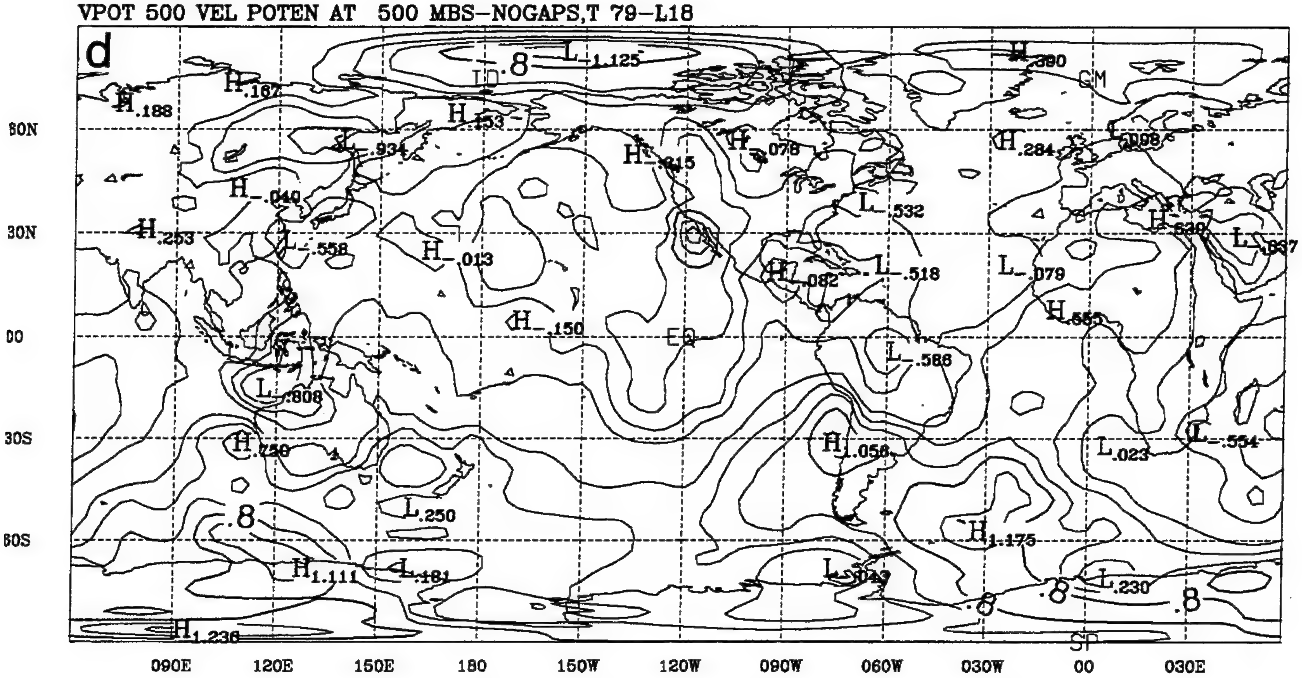

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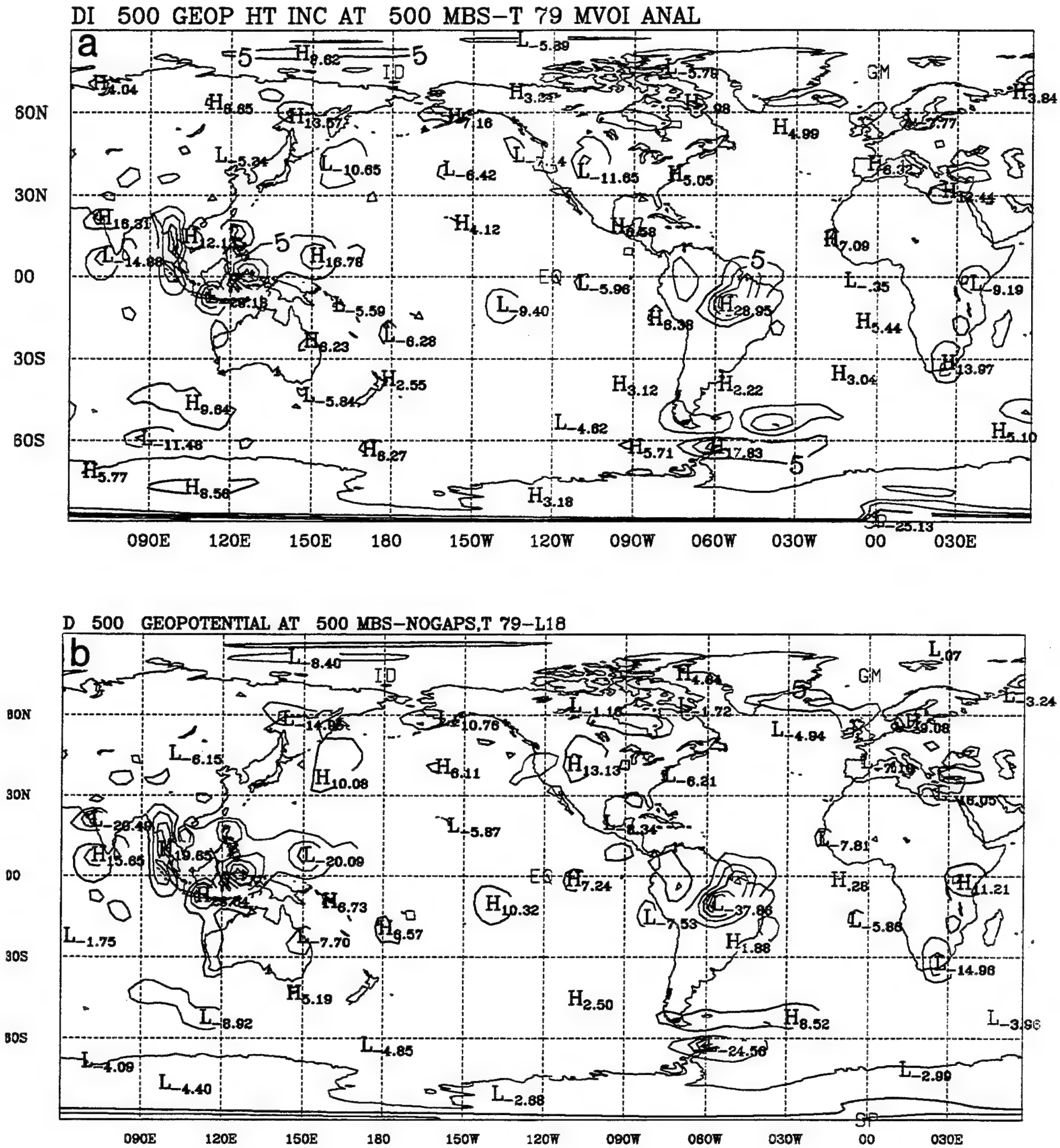

Figure 2. The 500-mb (a) height analysis increment, (b) height initialization increment, (c) velocity potential analysis increment, (d) velocity potential initialization increment at iteration 6 of analysis-initialization scheme for 0000 UTC 17 June 1995 (control experiment). Contour intervals are (a) and (b) 5 m, (c) and (d) 2×10^5 m^2 s^{-1}. The zero contour is omitted.

but is still quite large over much of the tropics. Some of the features in the tropical height analysis increment are actually larger at iteration 6 than at iteration 1, reflecting in part the vertical "spreading" of data rejection as discussed previously. There is also evidence that the scheme has begun to diverge by this point, as certain features (e.g., the center over Brazil) are strengthening with iteration number at all levels. In addition, the 500 mb height initialization increment (Fig. 2b) is consistently slightly larger in magnitude than the analysis increment; this likely results from a combination of both vertical "spreading" and divergence of the iterations. Certainly the scheme is nowhere near close to a satisfactory convergence, as low latitude height increments of 10–30 m are numerous in both Figs. 2a and 2b.

Figures 2c,d show the 500 mb velocity potential analysis and initialization increments for iteration 6. There is still cancellation between the two fields, but their magnitude has markedly decreased at all latitudes since iteration 1. This is particularly apparent over the Pacific ocean, among other regions. The maximum magnitude of velocity potential increment at iteration 6 is approximately half that at iteration 1. This decrease in intensity is observed at all levels and so is not due merely to a vertical redistribution of data rejection. Nevertheless, comparison of the sixth initialized analysis with the first uninitialized analysis (not shown) indicates significant differences; recall that the scheme is intended to minimize (to the extent possible) the difference between these two fields. Thus, although the velocity potential analysis is converging, it is to a state substantially different from the original analysis and so the method is not successful in retaining divergent wind information.

As a check on the generality of the preceding results, we repeated our control experiment with a second analysis valid 0000 UTC 8 July 1995. Results were generally quite similar, although the convergence of the velocity potential field was less uniform. Specifically, two features with rather large amplitude remained in the analysis and initialization increment even after six iterations; these were located over southeast Asia and northern South America. In other locations, however, the velocity potential behaved as in the previous experiment. The state towards which it converged was again significantly different from the original analysis.

In summary, results for this subsection are generally consistent with those of Van Tuyl

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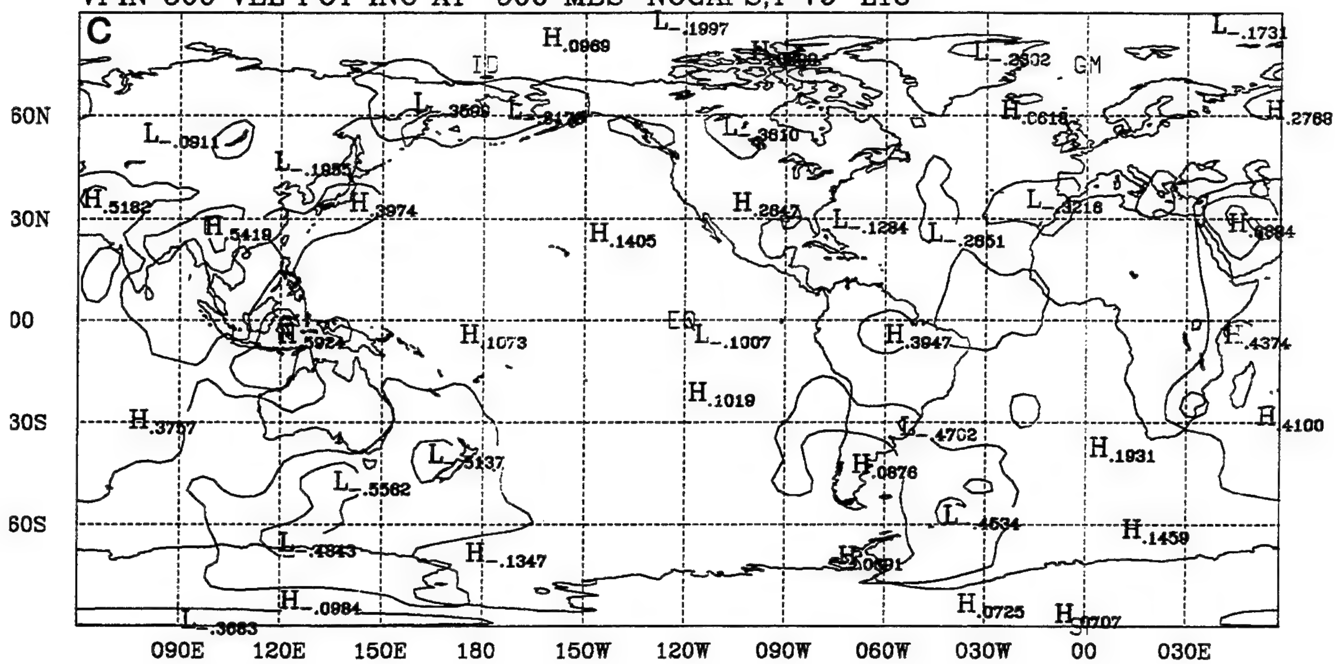

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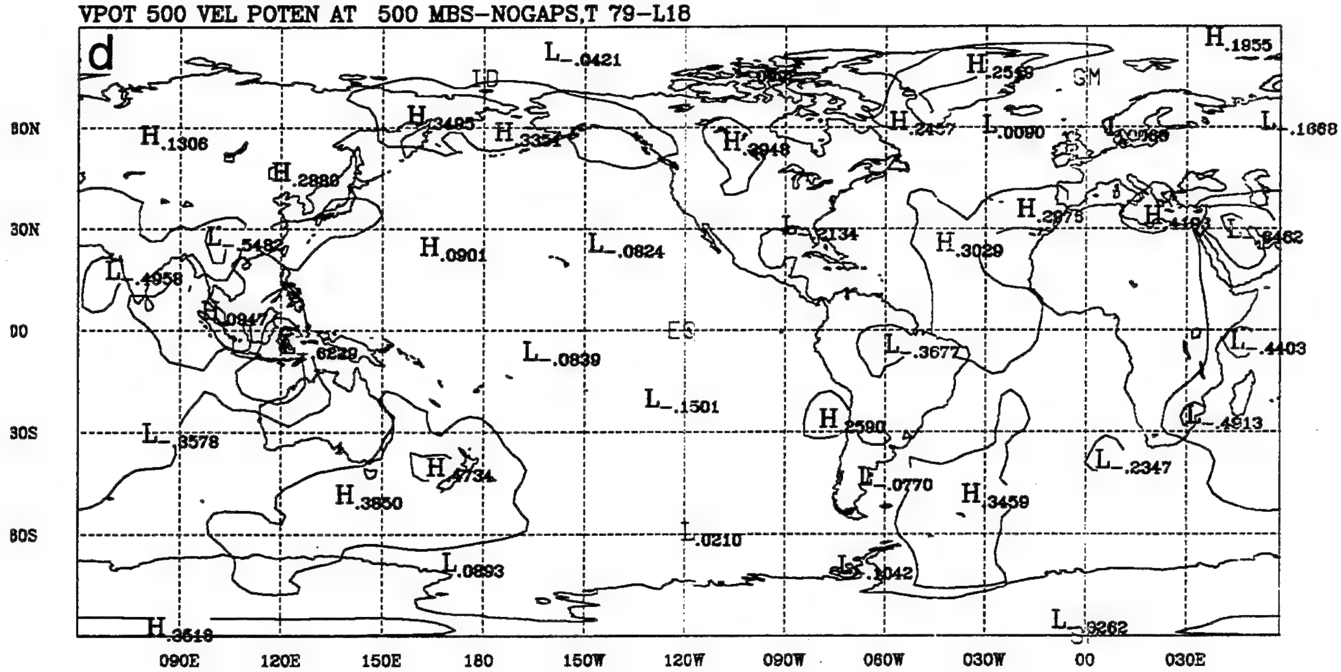

Figure 2 cont.

(1995), in that the combined analysis-initialization scheme achieves little if any improvement in the retention of low-latitude height and global divergent wind data. The primary difference with the shallow-water model results is that the velocity potential analysis and initialization increments decrease more strongly with iteration number in the present case. This is probably due to the ability of NOGAPS to represent the full three-dimensional structure of the divergence field, rather than just its vertical average, which tends to be a residual between cancelling positive and negative quantities. In the remainder of this study we examine the sensitivity of our results to changes in analysis parameter values, as well as to the inclusion of diabatic processes.

b. Sensitivity to Analysis Parameters

As in Van Tuyl (1995), we vary the analysis parameters in systematic fashion and determine the sensitivity, if any, of our results to these parameters. The specific parameters considered in this subsection are the geostrophic coupling parameter μ, the divergent coupling parameter ν, and the two SOAR parameters determining the decay scale and asymptote of the background error correlation function.

The tropical geostrophic coupling parameter (μ) was varied from its control case value 0.5 in increments of 0.25, following Van Tuyl (1995), resulting in a linear sequence $\mu = 0, 0.25, 0.5, 0.75, 1$. Figures 3a,b show the first 500 mb height initialization increment at 0000 UTC 17 June 1995 for, respectively, $\mu = 0.75$ and $\mu = 0.25$. The corresponding field for the control experiment appears in Fig. 1b. Comparison of the three initialization increments demonstrates that, as for the shallow-water case, the sensitivity of tropical data rejection to μ is quite small. There is a definite tendency for the height initialization increments (as well as analysis increments) at low latitudes to increase with decreasing μ; this result was also noted and explained by Van Tuyl (1995). (Certain features, for example those over Africa, change with μ in the opposite sense.) However, the maximum change between the $\mu = 0.25$ and $\mu = 0.75$ cases in Fig. 3 is only 1-2 m, demonstrating a lack of sensitivity to the geostrophic coupling. This conclusion is not affected by the inclusion of the $\mu = 0$ and $\mu = 1$ cases in the comparison. The above findings apply qualitatively at later iterations, although the magnitude of the difference between cases increases somewhat; consequently

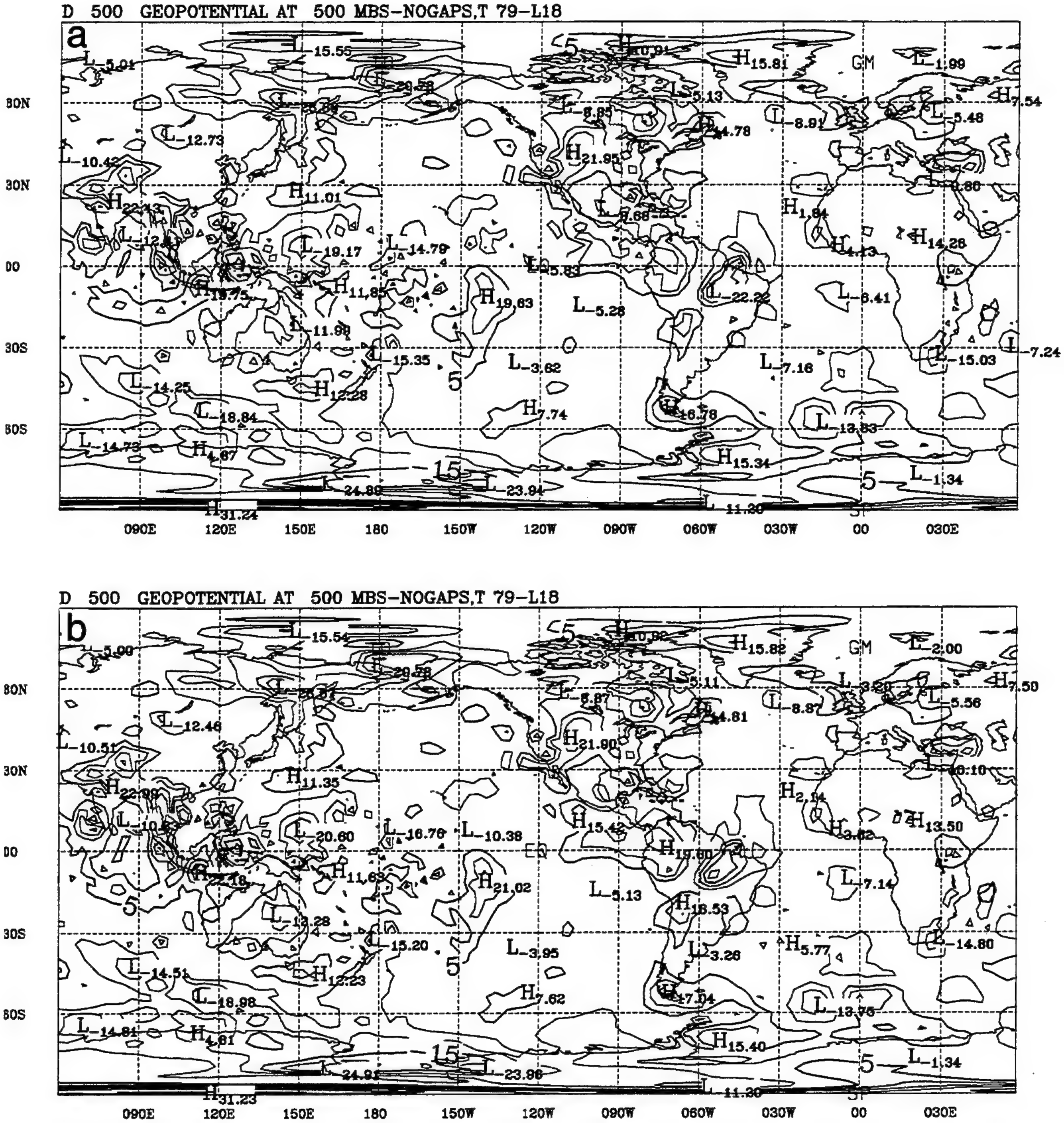

Figure 3. Height field initialization increment at 500 mb for tropical μ equal to (a) 0.75, (b) 0.25, at iteration 1 of analysis-initialization scheme for 0000 UTC 17 June 1995. Contour interval is 5 m. The zero contour is omitted. Analysis parameters are identical to those of control experiment except where noted.

the overall behavior of the scheme, in particular its convergence properties, is not strongly influenced by μ. A similar conclusion was reached in Van Tuyl (1995).

The value of μ may impact the divergent wind increment even though the form of the geostrophic coupling is nondivergent. This is because the constraint is applied individually in each analysis volume (refer to section 2b), but is not enforced on larger scales. An example of this influence occurs in the preceding series of experiments; the velocity potential increments (not shown) exhibit a feature over northern South America which intensifies strongly with increasing μ and is absent (or extremely weak) for $\mu = 0$. Since this feature occurs almost directly on the equator it is likely due to the change in sign of the Coriolis parameter between adjacent analysis volumes, and thus primarily spurious. For reasonable tropical values of μ the feature is small in magnitude, however, and should not significantly affect the analysis, especially since the initialization largely removes it. Nevertheless, this phenomenon illustrates a weakness of the volume method, and also demonstrates the problems which can arise from too strong a tropical geostrophic coupling. We note that all of our results regarding dependence on μ apply equally to the 0000 UTC 8 July 1995 experiment.

A sequence of experiments with tropical values of ν equal to 0,0.25,0.5,0.75,1 was performed; these were compared with each other and with the control case for which $\nu = 0.1$ at low latitudes. Results showed a rather weak dependence on ν, at least in terms of the performance of the iterative scheme. Height analysis and initialization increments tended to increase very slightly with increasing ν, as in the shallow-water experiments of Van Tuyl (1995); this phenomenon was related by Van Tuyl to an effective loosening of the geostrophic constraint. The velocity potential increments, as expected, increased in the tropics with increasing ν, becoming noticeably more noisy for $\nu \geq 0.5$. The convergence of the scheme was not qualitatively affected by ν for either height or velocity potential, however. As with our tropical μ experiments, results were virtually identical for both times examined (0000 UTC 17 June and 0000 UTC 8 July).

The first-guess error correlation function has two adjustable parameters which determine its decay scale and asymptote (section 3a). Two experiments were performed to test the sensitivity of our results to these parameters. In the first case the correlation length scale was doubled to 770 km and in the second the nondimensional asymptote was doubled to

0.2. Both of these increases have the effect of broadening the error correlation structure, which appears to possess a larger characteristic scale in the tropics than in midlatitudes (Thiébaux et al., 1986). Unfortunately, the NOGAPS optimum interpolation does not allow different correlation parameters for the midlatitude and tropical regions, so it was necessary to alter the correlation function globally. Analysis and initialization increments for height and velocity potential were larger in both experiments than in the control case, but this increase was significantly greater for the first experiment (in which the correlation length was doubled). Results for the second experiment showed only a very modest increase relative to control. The qualitative behavior of the scheme was not affected in either of the cases.

c. Nonconstant Background Error Variance

Thus far the background error field has been held constant, both in spatial structure and in magnitude, throughout the iterative scheme. This is a fairly severe restriction because the NOGAPS analysis assumes background error representative of a 6-hour forecast, which is not likely to be appropriate at later iterations. Van Tuyl (1995) and Williamson and Daley (1983) employed the zonally-averaged analysis error variance, obtained from the most recent optimum interpolation, to compute background error in certain of their experiments. This technique has serious difficulties, however. First, the analysis error variance is computationally rather time-consuming to determine, and is not typically calculated as part of an optimum interpolation. Second, this method still does not produce the complete analysis error *covariances* (or correlations), which must be modeled as before. Third, the analysis error variance determined from optimum interpolation applies to the *uninitialized* analysis, not the initialized analysis used as the background state; because nonlinear normal mode initialization is not a statistical technique, its effect on analysis errors is almost impossible to estimate. Finally, the analysis error is correlated with the error of the observations, violating the traditional assumption of optimum interpolation that observational and background errors are uncorrelated. (This assumption is of course violated by our scheme for any choice of background error, whether or not the violation is explicitly acknowledged in defining the error specification.)

Despite the above difficulties, one can at least argue qualitatively that the background

error variance should decrease with iteration number. The second background field (i.e., the first initialized analysis) contains the influence of the observations and is thus presumably more accurate than the first background. A similar argument holds for subsequent iterations. Because it is not clear how the error variance (let alone the covariance) should change quantitatively as a function of iteration number, we make the rather crude assumption that the variance decreases by a constant factor from one iteration to the next. Two experiments of this type were performed, differing only in the amount of decrease; specifically, the *variance* was decreased by either 0.75 or 0.5 between consecutive iterations (the standard deviations decreased by the square root of the appropriate factor). The usual (i.e., SOAR) correlation functions were employed, and all analysis parameters except error variance were held at their control case values.

Results for the two cases show weaker analysis and initialization increments relative to the control experiment, as expected; by the sixth iteration the increments in the second case (variance decrease of 0.5) are virtually zero. The tendency of the scheme to diverge for certain features (e.g. the height increment over Brazil) is also eliminated in both experiments. However, the final initialized state is not significantly altered, at least in the tropics, by the background error modification; Van Tuyl (1995) obtained a similar result. This lack of impact occurs because the analysis increments, particularly of height, are largely rejected at each iteration regardless of their magnitude, so that their cumulative effect remains small. Interestingly, a more substantial difference between cases occurs for the height field in midlatitudes, where little rejection exists; here, the repeated insertion of data evidently continues to influence the analysis even though optimum interpolation theory predicts that it should not. This result implies that our error statistics are considerably sub-optimal which, as previously discussed, is likely to be true after the first iteration at least.

d. Diabatic Initialization

Diabatic processes are a key component of the low-latitude dynamical balance, but are not typically included in nonlinear normal mode initialization. It is well known, however, that adiabatic Machenauer initialization tends to seriously weaken the analyzed Hadley circulation, because the convectively-forced divergence appears to the initialization as unbalanced gravity wave motion and is therefore eliminated (Puri and Bourke, 1982). Dia-

batic normal mode initialization was developed as a method of overcoming this and related problems; it typically employs the time-averaged, model-generated diabatic forcing as an additional (constant) term in the normal-mode balance (Wergen, 1988). Unfortunately, this technique also suffers from various shortcomings; for example, the model-generated diabatic processes, obtained from a short forecast, are usually not very accurate due to model "spin-up" and parameterization deficiencies. Even more seriously, the balance imposed by diabatic normal mode initialization is not generally appropriate for convectively-driven motions, particularly at low latitudes; refer to Errico (1984) and Errico and Rasch (1988). The proper inclusion of diabatic processes by initialization and data assimilation schemes represents a crucial issue in numerical weather prediction. Some of the more sophisticated techniques currently employed are described by Krishnamurti et al. (1991) and Puri and Davidson (1992); these involve the use of observed diabatic quantities, together with the full prediction model equations, to define the initial state.

Because NOGAPS possesses an option for diabatic normal mode initialization (although it is not used operationally), we performed two experiments, one for each analysis time, in which this scheme was employed in the initialization step. In all other respects these experiments are identical to the two control cases discussed in section 3a. Figures 4a,b present the 500 mb height initialization increments at iteration 1 and iteration 6, respectively, for the 0000 UTC 17 June 1995 case with diabatic initialization. (All results for the corresponding 0000 UTC 8 July 1995 case are qualitatively identical.) Comparing Fig. 4 and Fig. 1, we observe that the effect of including diabatic initialization is to slightly decrease the magnitude of the initialization increments, indicating less rejection. However, this decrease is not uniform (some features exhibit an increase in magnitude), and the overall impact on rejection is extremely small. Similar results apply to the height field at other levels, as well as velocity potential. These findings are not surprising, because the vertical modes initialized in NOGAPS are essentially barotropic (in the troposphere) and thus not strongly affected by diabatic forcing. (Refer to comments in section 2c.) As expected, comparison of the final initialized analysis with and without diabatic initialization (not shown) also indicates little difference. It is conceivable, of course, that diabatic effects would produce greater impact if higher vertical modes were included in the initialization, but we did not investigate this

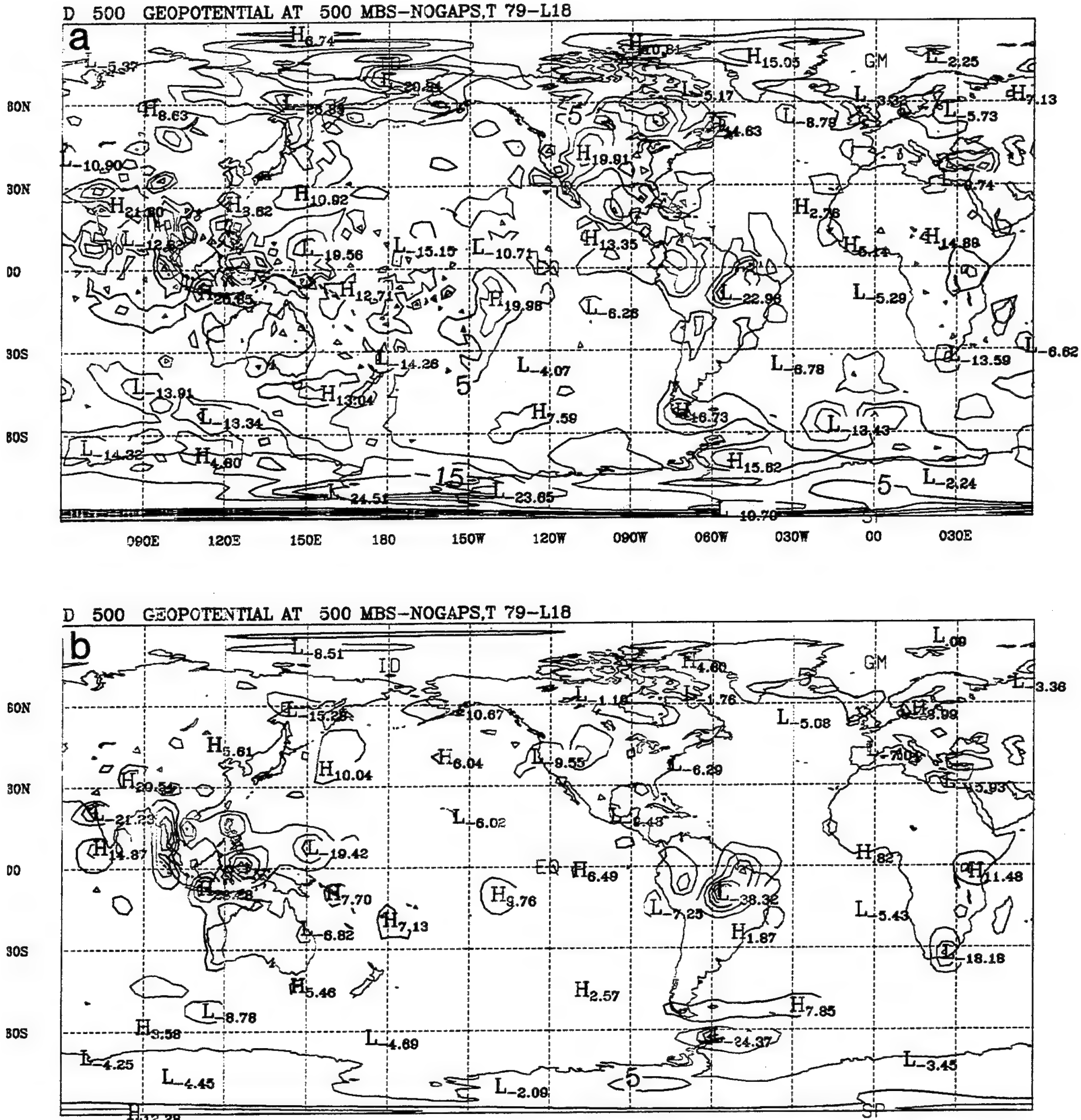

Figure 4. The 500-mb height field initialization increment at (a) iteration 1, (b) iteration 6, of analysis-initialization scheme with diabatic normal mode initialization for 0000 UTC 17 June 1995. Contour interval is 5 m. The zero contour is omitted. Analysis parameters are identical to those of control experiment.

possibility.

4. Summary and Conclusions

This study was designed to extend the shallow-water model results of Van Tuyl (1995) to a multilevel, operational model (NOGAPS). Despite the use of a much more sophisticated model, the basic finding remained that the iterative scheme had only slight positive impact (at best) on the retention of low-latitude height and global divergent wind data. Thus, the failure of this technique in Van Tuyl (1995) was not due to a deficiency of the shallow-water model, or to the use of synthetic data interpolated from an analysis. The most significant difference with Van Tuyl's results was the improved convergence of the velocity potential analysis and initialization increments in NOGAPS, that is, the increments diminished more rapidly with iteration number. This was probably due to the presence of the full three-dimensional divergence in the multilevel model, as opposed to the vertically-integrated divergence employed by the shallow-water system. However, the final initialized state was still significantly different from the original analysis, and thus little net improvement in divergence retention was achieved.

The performance of the scheme was not sensitive to the values of any of the optimum interpolation parameters, in agreement with the results of Van Tuyl (1995). Increasing the low-latitude geostrophic coupling parameter did slightly diminish the magnitude of the height increments, but too large a value produced undesirable features in the tropical velocity potential field. The divergent coupling at low latitudes had minimal impact until it was increased to a rather high value; then, its main effect was to produce "noise" in the velocity potential increments. Decreasing the background error variance at each iteration significantly reduced the magnitude of the analysis and initialization increments, but had little effect on the final initialized state in the tropics. (A somewhat greater impact did occur in midlatitudes, however.)

The use of diabatic normal mode initialization tended to very slightly decrease data rejection; however, the effect was minimal because the initialized modes were not those stongly influenced by diabatic processes. It is possible that including more vertical modes, or changing the frequency cutoff in the Machenauer scheme, would have produced significantly

different results. This was not attempted, however, because of the likelihood that the initialization would diverge, and also because of the general inappropriateness of the diabatic Machenauer balance.

Although the rejection of low latitude height and divergent wind information is certainly not desirable, as it indicates deficiencies in either the analysis or initialization (or both), its practical significance in terms of operational numerical weather prediction is probably minimal. This is because, as pointed out by Van Tuyl (1995), the rejected variables are those which are less reliably observed in the tropics. The height increment features, being limited to isolated regions in the vicinity of radiosonde stations, are especially suspect. Certain of these stations have in fact been "blacklisted" operationally due to unusually large observation minus first guess differences (N. Baker, personal communication). Of course, rejection of questionable data by the initialization scheme is not an ideal solution, in part because it is a function of latitude (height data is not rejected outside of the tropics), and also because it cannot account for data quality. At a given latitude, and for a given degree of imbalance, good and poor height data will be rejected equally. It is essentially fortuitous that height rejection is maximized where the data tends to be least reliable.

The data rejection in NOGAPS (and in the shallow-water model) is due primarily to the great difference in dynamical balance between the analysis and initialization, at least at low latitudes. As discussed by Van Tuyl (1995), the optimum interpolation produces a mass-wind balance in the tropics substantially different from that of nonlinear normal mode initialization; thus, either the analyzed height or wind must be rejected to satisfy the normal mode condition. The failure of the iterative scheme for a variety of analysis parameters, as well as for a diabatic normal mode initialization, suggests that this rejection problem is inherent in the analysis and initialization techniques themselves, i.e. optimum interpolation and nonlinear normal mode initalization. Thus, a resolution must await the implementation of improved data assimilation methods.

Current analysis and initialization schemes in operational use generally neglect diabatic processes, which are a vital part of the low-latitude mass-wind balance. As discussed in section 2, optimum interpolation essentially decouples the mass and wind in the tropics, and it is incapable of incorporating diabatic effects (or any nonlinearity) in its dynamical

balance. Nonlinear normal mode initialization also cannot include diabatic forcing in a realistic manner (refer to section 3d). Physical initialization schemes (e.g. Krishnamurti et al., 1991) are likely to produce significant improvements in this respect. Other techniques such as three-dimensional variational assimilation (Parrish and Derber, 1992) and four-dimensional assimilation should also prove useful in obtaining more accurate and better-balanced tropical analyses. One advantage of three-dimensional assimilation, for example, is that it can use all of the available data at once; thus, problems associated with the overlapping volumes of optimum interpolation (refer to section 3b) are eliminated. This method can also directly assimilate satellite radiances (Parrish and Derber, 1992), possibly resulting in a better estimate of the tropical mass field. As these and other new techniques are introduced operationally, the data rejection problem is likely to disappear as part of the overall advance of data assimilation.

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References

Baker, N. L., 1992: Quality control for the Navy operational atmospheric database. *Wea. Forecasting*, **7**, 250-261.

Ballish, B. A., 1981: A simple test of the initialization of gravity modes. *Mon. Wea. Rev.*, **109**, 1318-1321.

Barker, E. H., 1992: Design of the Navy's multivariate optimum interpolation analysis system. *Wea. Forecasting*, **7**, 220-231.

Daley, R., 1978: Variational nonlinear normal mode initialization. *Tellus*, **30**, 201-218.

—, 1980: On the optimal specification of the initial state for deterministic forecasting. *Mon. Wea. Rev.*, **108**, 1719-1735.

—, 1985: The analysis of synoptic scale divergence by a statistical interpolation procedure. *Mon. Wea. Rev.*, **113**, 1066-1079.

Davidson, N. E., and K. Puri, 1992: Tropical prediction using dynamical nudging, satellite-defined convective heat sources, and a cyclone bogus. *Mon. Wea. Rev.*, **120**, 2501-2522.

Errico, R. M., 1983: Convergence properties of Machenauer's initialization scheme. *Mon. Wea. Rev.*, **111**, 2214-2223.

—, 1984: The dynamical balance of a general circulation model. *Mon. Wea. Rev.*, **112**, 2439-2454.

—, and P. J. Rasch, 1988: A comparison of various normal-mode initialization schemes and the inclusion of diabatic processes. *Tellus*, **40A**, 1-25.

—, T. E. Rosmond, and J. S. Goerss, 1993: A comparison of analysis and initialization increments in an operational data-assimilation system. *Mon. Wea. Rev.*, **121**, 579-588.

Fillion, L., 1991: Variational implicit normal mode initialization on the sphere. *Mon. Wea. Rev.*, **119**, 631-652.

—, and M. Roch, 1992: Variational implicit normal-mode initialization for a multilevel model. *Mon. Wea. Rev.*, **120**, 1050-1076.

Franke, R., E. Barker, and J. Goerss, 1988: The use of observed data for the initial-value problem in numerical weather prediction. *Comput. Math. Applic.*, **16**, 169-184.

Gandin, L. S., 1963: Objective analysis of meteorological fields. *Gidrometeorologicheskoe Izdatelstvo*, Leningrad. Translated from Russian, Israeli Program for Scientific Translations, Jerusalem, 1965, 242pp.

Goerss, J. S., and P. A. Phoebus, 1992: The Navy's operational atmospheric analysis. *Wea. Forecasting*, **7**, 232-249.

Hogan, T. F., and T. E. Rosmond, 1991: The description of the Navy Operational Global Atmospheric Prediction System's spectral forecast model. *Mon. Wea. Rev.*, **119**, 1786-1815.

Krishnamurti, T. N., J. Xue, H. S. Bedi, K. Ingles, and D. Oosterhoff, 1991: Physical initialization for numerical weather prediction over the tropics. *Tellus*, **43AB**, 53-81.

Lorenc, A. C., 1981: A global three-dimensional multivariate statistical interpolation scheme. *Mon. Wea. Rev.*, **109**, 701-721.

Machenauer, B., 1977: On the dynamics of gravity oscillations in a shallow water model, with applications to normal mode initialization. *Contrib. Atmos. Phys.*, **50**, 253-271.

Parrish, D. F., and Derber, J. C., 1992: The National Meteorological Center's spectral statistical-interpolation analysis system. *Mon. Wea. Rev.*, **120**, 1747-1763.

Puri, K., and W. Bourke, 1982: A scheme to retain the Hadley circulation during nonlinear normal mode initialization. *Mon. Wea. Rev.*, **110**, 327-335.

Rossby, C.-G., 1938: On the mutual adjustment of pressure and velocity distributions in certain simple current systems, II. *J. Marine Res.*, **1**, 239-263.

Temperton, C., 1989: Implicit normal mode initialization for spectral models. *Mon. Wea. Rev.*, **117**, 436-451.

Thiébaux, H. J., H. L. Mitchell, and D. W. Shantz, 1986: Horizontal structure of hemispheric forecast error correlations for geopotential and temperature. *Mon. Wea. Rev.*, **114**, 1048-1066.

Tribbia, J. J., 1982: On variational normal mode initialization. *Mon. Wea. Rev.*, **110**, 455-470.

Unden, P., 1989: Tropical data assimilation and analysis of divergence. *Mon. Wea. Rev.*, **117**, 2495-2517.

Van Tuyl, A. H., 1995: Evaluation of a scheme for improving the acceptance of low-latitude data in atmospheric data assimilation. *Mon. Wea. Rev.*, **123**, 2515-2533.

Wergen, W., 1988: The diabatic ECMWF normal mode initialization scheme. *Contrib. Atmos. Phys.*, **61**, 274-302.

Williamson, D. L., R. Daley, and T. W. Schlatter, 1981: The balance between mass and wind fields resulting from multivariate optimal interpolation. *Mon. Wea. Rev.*, **109**, 2357-2376.

—, and —, 1983: A unified analysis-initialization technique. *Mon. Wea. Rev.*, **111**, 1517-1536.